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FINAL TECHNICAL REPORT FOR OFFICE OF NAVAL OCEANOGRAPHY GRANT #N00014-90-J-1268

"Modeling Studies of Ice-Ocean Dynamics and Thermodynamics"

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March 10, 1993

File: 343-6055

FINAL TECHNICAL REPORT

"MODELING STUDIES OF ICE-OCEAN DYNAMICS AND THERMODYNAMICS"

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George L. Mellor, Principal Investigator
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This report covers the period, October 1, 1989 to September 30, 1992. It was preceded by contract number N00014-84-K-0640 for which we previously reported completion of eight journal articles. These papers were mainly concerned with the development of an ice model and the coupling of that model to the (now standard) Princeton ocean model. The studies were largely based on either one-dimensional, time dependent models or two-dimensional (x,z) models. The latter included a model study of atmospheric-ice interactions.

The more current grant produced four papers. The first, "One hundred years of Arctic ice cover variations as similated by a one-dimensional ice-ocean model" was an application of our time dependent, one-dimensional model to climate time scale variability. The second paper "An equation of state for numericla models of oceans and estuaries" was a modification of the UNESCO equation of state which rendered it useful for numerical ocean models.

The next paper, "Modeling deep convection in the Greenland Sea" represented the emergence of our fully three-dimensional, ice-ocean coupled model. Here it was applied to the process study of deep convection and bottom water formation. Finally, the paper "Modeling the seasonal variability of a coupled Arctic ice-ocean system" was a modeling study of the Arctic Ocean and the Norwegian-Greenland-Barents Seas and included multi-year calculations and sensitivity studies. The first page from these papers are attached to this report.

It should be noted that this grant and its predecessors contributed to general improvement of the Princeton Ocean Model which, through an electronic users group, is now in use throughout the world; this includes Naveoceano and other Navy sponsored research projects.

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 95, NO. C9, PAGES 15,959–15,969, SEPTEMBER 15, 1990

One Hundred Years of Arctic Ice Cover Variations as Simulated by a One-Dimensional, Ice-Ocean Model

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A one-dimensional ice-ocean model consisting of a second moment, turbulent closure, mixed layer model and a three-layer snow-ice model has been applied to the simulation of Arctic ice mass and mixed layer properties. The results for the climatological seasonal cycle are discussed first and include the salt and heat balance in the upper ocean. The coupled model is then applied to the period 1880–1985, using the surface air temperature fluctuations from Hansen et al. (1983) and from Wigley et al. (1981). The analysis of the simulated large variations of the Arctic ice mass during this period (with similar changes in the mixed layer salinity) shows that the variability in the summer melt determines to a high degree the variability in the average ice thickness. The annual oceanic heat flux from the deep ocean and the maximum freezing rate and associated nearly constant minimum surface salinity flux did not vary significantly interannually. This also implies that the oceanic influence on the Arctic ice mass is minimal for the range of atmospheric variability tested.

1. INTRODUCTION

The main goal of this paper is to explore the behavior of the ice–upper ocean system on climatic time scales and especially the asymmetry of the freezing and melting seasons. We study here only the Arctic ice cover, which is a prominent climate indicator since it contains a large multiyear ice component. Also, a goal is to further test the one-dimensional ice-ocean model of *Mellor and Kantha* [1989] (hereinafter referred to as MK), in which a change in a boundary condition at the ice-ocean interface has been implemented.

The influence of the polar ice cover on climate derives mainly from the fact that the ice cover can insulate the ocean and thereby modify the surface heat flux by an order of magnitude. The insulating effect depends on the reflective properties of the ice surface, the ice concentration, and the ice thickness [*Maykut*, 1978, 1982; *Maykut and Untersteiner*, 1971]. Climate models are shown to have a strong sensitivity to snow and ice cover albedo in Polar regions [*Manabe and Stouffer*, 1980].

Leads are very important to climate response, since they release heat to the atmosphere in the wintertime and, as a secondary effect, lower the surface albedo. Heat loss through a 1% lead area can be as large as the conductive loss through the remaining 99% of the area covered by thick ice. The estimates for lead area range from 1 to 5% for winter months. Variability in the lead area has been shown to affect considerably the surface air temperatures in climate models [*Ledley*, 1988*a*, *b*].

Monitoring the changes in ice cover have been limited to records of ice extent and concentration in the marginal ice zones before the advent of satellites in the 1970s [*Walsh and Johnson*, 1979] or to a local ice index such as compiled for the Icelandic coastal waters [*Sigtryggson*, 1972]. The latter record beginning in 1900 shows that there are periods with extremely heavy ice around 1910 and in the end of the 1960s. Remote sensing techniques have now made it possible to monitor the ice concentration changes from space (e.g., the passive microwave sensors such as electrically scanning microwave radiometer (ESMR), [*Parkinson et al.*, 1987], scanning multichannel microwave radiometer (SMMR), or special sensor microwave/imager (SSM/I)). The ice thickness variations pose a much more difficult monitoring task. Even though SMMR and SSM/I can resolve thin ice and multiyear ice components of the ice concentration, they cannot measure actual thickness of the sea ice. Upward looking sonar along submarine tracks has been the only measurement able to give a large-scale survey of the ice thickness variability in the Arctic [*Bourke and Garrett*, 1987; *Wadhams*, 1988].

On climatic time scales the ice thickness is an important factor in controlling the conductive heat flux through ice and thus the salinity flux to the ocean arising from bottom ice accretion. Because ice stores and releases fresh water, the ice thickness and concentration variations, which determine the ice mass, are important quantities in the hydrological cycle. (Note, however that the freshwater reservoir of Arctic ice is a thousand times smaller than that of the Greenland ice sheet [*Goody*, 1980]. This sea ice–fresh water source can affect the subpolar gyre, mainly through the ice exported from the Arctic to the Greenland Sea, and can significantly modify the oceanic surface layers in the areas conducive to water mass modification such as in the Greenland, Iceland, and Labrador seas [*Swift and Aagaard*, 1981].

The ice export from the Arctic through the Fram Strait depends also on the winds and ocean currents in the outflow region. Estimates of yearly averaged areal outflows are 0.9 × 10^6 km^2 [*Vowinckel*, 1964], 0.84 × 10^6 km^2 [*Moritz*, 1988], and 1.55 × 10^6 km^2 [*Vinje and Finnekåsa*, 1986]. Large-scale ice-ocean models have predicted the ice export to vary up to a factor of 3 interannually, with a mean of 1400 km^3 for volumetric and a mean of 0.63 × 10^6 km^2 for the areal export [*Walsh et al.*, 1985]. The ice export variability in the model of Walsh et al. is very sensitive to the wind forcing, which contributes 75% to the ice drift. This is in contrast to estimates where wind and current drift portions are 30% and 70%, respectively, as derived from observed buoy drifts and geostrophic winds [*Moritz*, 1988]. Empirical models for the ice outflow based on the air pressure difference across the Fram Strait suggest that the wind-driven flow may have

Paper number 90JC00754.
0148-0227/90/90JC-00754$05.00

Reprinted from JOURNAL OF ATMOSPHERIC AND OCEANIC TECHNOLOGY, Vol. 8, No. 4, August 1991
American Meteorological Society

An Equation of State for Numerical Models of Oceans and Estuaries

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26 October 1990 and 3 January 1991

1. Introduction

Numerical models of oceans or estuaries require an equation of state in order to relate density, salinity, temperature, and pressure. The needs are for an equation that is reasonably efficient numerically and has a wide range of application. In addition, since numerical models solve for potential temperature, one would like an equation of state in the form

$$\rho = \rho(S, \Theta, p) \qquad (1)$$

where ρ is density, S is salinity, Θ is potential temperature, and p is pressure.

In the following we will denote property functions as in Eq. (1) where, for example, $\partial\rho/\partial S$ denotes the partial derivative of density with respect to S holding θ and p constant. We denote spatial functions with a superposed tilde so that, for example, $\partial\tilde{\rho}/\partial x$ denotes the partial derivative of density with respect to x holding y and z constant.

Models require horizontal density gradients, $(\partial\tilde{\rho}/\partial x, \partial\tilde{\rho}/\partial y)$, which are readily obtained from Eq. (1) and known spatial fields of $\tilde{T}$, $\tilde{S}$, and $\tilde{p}$. They also require the vertical static stability obtained either from

$$\frac{N^2}{g} = -\frac{1}{\rho}\left[\frac{\partial\rho}{\partial S}\frac{\partial\tilde{S}}{\partial z} + \frac{\partial\rho}{\partial\theta}\frac{\partial\tilde{\theta}}{\partial z}\right] \qquad (2a)$$

or from

$$\frac{N^2}{g} = -\frac{1}{\rho}\left[\frac{\partial\tilde{\rho}}{\partial z} - \frac{1}{c_s^2}\frac{\partial\tilde{p}}{\partial z}\right] \qquad (2b)$$

where $c_s^2 = [(\partial\rho/\partial p)_{s,\theta}]^{-1}$ is the speed of sound squared. The term N^2 governs vertical mixing either through a Richardson number formulation (e.g., Munk and Anderson 1948) or by a turbulence closure scheme (e.g., Mellor and Yamada 1982). Equations (2a) and (2b) are equivalent since $\partial\tilde{\rho}/\partial z = (\partial\rho/\partial S)(\partial\tilde{S}/\partial z) + (\partial\rho/\partial\theta)(\partial\tilde{\theta}/\partial z) + (\partial\rho/\partial p)(\partial\tilde{p}/\partial z)$. However, numerical calculations using (2a) or (2b) may differ as discussed below.

Bryan and Cox (1972) use a three-dimensional model whose vertical coordinate corresponds to constant depth levels. To obtain density they preprocess a table of coefficients for each depth. On the other hand, we have a three-dimensional, primitive-equation ocean model that uses a sigma coordinate system in the vertical (Oey et al. 1986a,b; Blumberg and Mellor 1987). It has been applied to coastal oceans and estuaries but, increasingly, is also applied to deep water ocean domains. Other sigma coordinate models have appeared and are still appearing in the literature. For sigma coordinate models, the Bryan–Cox scheme is inappropriate.

2. Analysis

The now standard, UNESCO (United Nations Educational, Scientific, and Cultural Organization) equation of state is fairly expensive computationally and is overly precise relative to the capability of numerical models to produce comparable precision. Furthermore, it is an equation of the form, $\rho = \rho(S, T, p)$; i.e., in situ temperature is an independent variable rather than potential temperature.

A less precise but substantially simpler formula of the desired form is

$$\rho(S, \Theta, p) = \rho(S, \Theta, 0) + \frac{p}{c^2}\left(1 - C\frac{p}{c^2}\right) \qquad (3)$$

where $\rho(S, \Theta, 0)$ can be taken from the UNESCO formula [the complete formula, $\rho(S, T, p)$, increases the computational time by about a factor of 3] since $\Theta = T$ for $p = 0$. The second term in Eq. (3), the pressure dependent part, contains $c = c(S, \Theta, p)$ and a constant C. It should be noted that c is not exactly the speed of sound. Thus, from Eq. (3)

$$c_s^2 = \frac{1}{(\partial\rho/\partial p)_{s,\theta}} = c^2\left[\left(1 - \frac{2p}{c}\frac{\partial c}{\partial p}\right)\left(1 - 2C\frac{p}{c^2}\right)\right]^{-1}. \qquad (4)$$

To approximate the complete UNESCO function, we have determined that

$$\rho(S, \Theta, p) = \rho(S, \Theta, 0) + 10^4\frac{p}{c^2}\left(1 - 0.20\frac{p}{c^2}\right) \qquad (5a)$$

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JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 97, NO. C4, PAGES 5389–5408, APRIL 15, 1992

Modeling Deep Convection in the Greenland Sea

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The development of deep convective events in the high-latitude ocean is studied using a three-dimensional, coupled ice-ocean model. Oceanic mixing is described according to the level 2.5 turbulence closure scheme in which convection occurs in a continuous way, i.e., convective adjustment is not invoked. The model is forced by strong winds and surface cooling. Strong upwelling at the multiyear ice edge and consequent entrainment of warm Atlantic waters into the mixed layer is produced by winds parallel to the ice edge. Concomitant cooling drives deep convection and produces chimneylike structures. Inclusion of a barotropic mean flow over topography to the model provides important preconditioning and selects the location of deep convection. The most efficient preconditioning occurs at locations where the flow ascends a slope. In a stratified environment similar to the Greenland Sea with a 12 m s^{-1} wind the model simulations show that localized deep convection takes place after about 10 days to depths of 1000 m.

1. INTRODUCTION

Deep water formation at high latitudes in the world oceans provides cold water masses which spread equatorward and balance the advection of the warm surface waters from the low latitudes. Deep water formation has long been thought to occur on the continental shelves around Antarctica and the Arctic Ocean as a result of the brine-enriched cold shelf waters sinking to the deep basins. However, it has not been established how deep and on what time scale this sinking occurs. The geostrophic constraint will force the heavy waters to follow the coast, and only the frictional forces at the bottom can cause cross-isopycnal flow. Tracer studies in the Arctic show that the deep waters in the Canadian basin are at least 700 years old [*Östlund et al.*, 1987], which means that the cold, brine-enriched shelf waters from the vast East Siberian shelves have not been sinking to the deeper basin. On the other hand, the data from the Eurasian side of the Arctic show that in the Nansen basin, the deep waters are only a few decades old [*Wallace and Krysell*, 1987]. *Östlund et al.* [1987] give a more specific age of 30 years for the Eurasian deep water.

Another deep water source is open ocean production via deep convection, most notably to the Greenland and Iceland seas in the northern hemisphere and to the Weddell Sea in the southern hemisphere. In the southern hemisphere the manifestation of a convection event, as a deep, nearly uniformly stratified water column beneath the summer mixed layer, has been found in the Weddell Sea by *Gordon* [1978]. This Weddell Sea "chimney" was about 15 km in diameter, about 4–5 times the typical baroclinic Rossby radius of deformation in the area. There is only one known deep convection area in the temperate zone, namely, the Gulf of Lions in the Mediterranean, where the cold northerly Mistral and Tramontaine winds cause strong mixing and deep convection in the cyclonic gyre located in the area [*MEDOC Group*, 1970]. Deep convection extending to 1500 m can also occur in the Labrador Sea [*Clarke and Gascard*, 1983] where this process is responsible for the formation of the Labrador Sea water, an important component of the Atlantic Intermediate water [*McCartney and Talley*, 1984].

Investigators searching for chimneys in the Greenland Sea were unable to find these mesoscale features until very recent years, when field programs such as the Marginal Ice Zone Experiment (MIZEX) and the Greenland Sea project were conducted. During the winter MIZEX, some convective areas, about 10 km across, were found near the ice edge region [*MIZEX '87 Group*, 1989]. These chimneys extended at least to 250–300 m, which was the depth limit of the (towed Sea Soar) hydrographic survey. The temperatures and salinities in the chimneys ranged from 0° to 0.5°C and from 34.84 to 34.94 ppt respectively. *Rudels et al.* [1989] reported finding a convective area with a depth of 1250 m and an unknown spatial extent from a 1987 winter cruise to the Greenland Sea. *Bogorodsky et al.* [1987] have reported convective areas, about 100 km across, in the Greenland Sea which extended to the bottom depth of 3500 m. *Johannessen et al.* [1990] have found further evidence of wintertime convection in the Boreas basin, the northern most part of the Greenland Sea. In all of the aforementioned studies the information of the horizontal extent of the chimneys beyond the observed cross section is uncertain. Based on the observations of *Johannessen et al.* [1990], some of the "chimneys" may actually be elongated features, even though they suggest that chimneys originate from cyclonic eddies which are abundant in that particular area.

On the basis of the hydrographic data of the deep convective area by *Gordon* [1978], *Killworth* [1979] used a nonpenetrative convection model to study possible mechanisms and necessary surface forcing to produce these features which he termed "chimneys." He concluded that there must be some preconditioning mechanism which selected a particular overturning area. As a possible mechanism he suggested baroclinic instability; eddies resulting from the instability occur at the same spatial scales, and the pycnocline in the cyclonic eddies is nearer to the surface than the surrounding water, thus giving rise to enhanced mixing.

Martinson et al. [1981] later revised the model of *Killworth* [1979] to study deep convection in the large polynyas occur-

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Paper number 92JC00061.
0148-0227/92/92JC-00061$05.00

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 97, NO. C12, PAGES 20,285–20,304, DECEMBER 15, 1992

Modeling the Seasonal Variability of a Coupled Arctic Ice-Ocean System

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Results from modeling studies of the ice-ocean system in the Arctic Basin and in the Norwegian-Greenland-Barents seas are presented. We used a three-dimensional coupled ice-ocean model developed at Princeton University. The ocean model applies the primitive equations and a second moment turbulence closure for turbulent mixing. The snow-ice model uses a three-level thermodynamic scheme which resembles Semtner's (1976*a*) model. Our conclusions based on the seasonal simulations are as follows. (1) Using monthly climatological surface heat flux and wind stress, the seasonal variability of the ice cover is quite realistic in that the thickest ice is located north of Greenland and the average ice thickness is about 3 m. The largest deviation between the simulated and observed ice cover is in the Greenland Sea where oceanic conditions determine the ice edge. Basically, the monthly climatological forcing does not result in strong enough mixing to bring sufficient heat from the deep ocean to keep the central Greenland Sea gyre ice free. The results improve for both the ice cover and ocean by invoking daily wind forcing for which we first chose year 1987. In the ocean model, the large mixing events associated with storm passages are resolved, and as a result, the overall oceanic structure in the Greenland Sea appears to be more realistic. However, no deep convection takes place in the model during 1987 which is likely the result of diminished storm activity in the northern part of the Greenland Sea. The ice thickness field appears to be very anomalous 1987, so an experiment with 1986 daily wind forcing was also done, which resulted in an ice thickness field similar to some reported from other ice models. (2) Both monthly and daily surface forcing result in a similar behavior of the Atlantic waters in the Arctic Basin. The Atlantic waters circulate at about the observed level, between 400 and 600 m. The survival of the Atlantic waters in the basin depends strongly on the heat loss through the ice cover, and it appears that too much heat is lost on the Eurasian side through the ice because the simulated Atlantic waters are too cool by about 0.2–0.5°C. (3) For the monthly climatology case, a large amount of cold and salty water enters the Eurasia Basin from the Kara and Laptev seas area and finds its way toward the Canada Basin. This water mass appears to result from ice formation in the Kara and Laptev seas. When applying the daily forcing, this deep salinity maximum disappears due to increased mixing on the shelves. Nevertheless, this suggests a mechanism within the Arctic Ocean as to why the deep Canada Basin is much saltier than the Eurasia Basin.

1. INTRODUCTION

The objective of this study is to explore the circulation of ice and water masses in the Arctic Ocean and its peripheral seas, and the role of ice thermodynamics-dynamics in these processes. The Arctic Ocean, together with Greenland, Iceland, Norwegian, and Barents seas, comprise a unique system where a large portion of the world ocean deep waters are formed. Inflows and outflows in this area such as river runoff, Atlantic and Pacific inflow, and ice export determine the overall stratification structure. The freshwater component is especially important because of its stabilizing effect; in the Arctic Ocean, it prevents heat exchange between upper and deeper parts of the water column; in the Greenland Sea, an excess freshwater cap in the form of ice can prohibit the renewal of deep waters.

The Greenland, Iceland, Norwegian, and Barents seas form a gateway to the Arctic, where the Atlantic and Arctic origin waters mix and form new water masses through convective processes. The East Greenland Current represents the major outflow from the Arctic Basin and it is a boundary current closely following the continental slope. The Atlantic waters enter the area through the Faeroe-Shetland Channel and flow northward along the Norwegian margin. Part of this current branches off into the Barents Sea. The main body of the Norwegian Current, however, continues northward west of Svalbard, where it mixes with, and plunges beneath, the Arctic surface waters. This deep current forms a counterclockwise flow in the Eurasia Basin. The same water mass also has an entrained westward branch which forms a large cyclonic gyre in the Greenland Sea. A schemetic picture of the different surface currents is shown in Figure 1. A major subsurface outflow from the Greenland-Iceland-Norwegian seas into the Atlantic occurs in the Denmark Strait and across the ridges between Iceland and Scotland. This overflow has been the subject of intensive investigations sponsored by the International Council for the Exploration of the Sea (ICES) [*Johannessen*, 1986; *Swift*, 1986].

Compared to the Greenland Sea our knowledge of the Arctic circulation is more sketchy, especially for the subsurface. The Arctic surface water circulation is reasonably well known and can be monitored by the drift patterns of sea ice [*Thorndike and Colony*, 1982]. The surface water in the Canada Basin, as a long-term mean, moves in an anticyclonic gyre, while the Eurasia Basin is dominated by the Transpolar Current [*Gordienko and Laktionov*, 1969, Figure 1], originating in the Chukchi Sea and flowing out in Fram Strait. The circulation of the mixed layer, however, does not

Paper number 92JC02037.
0148-0227/92/92JC-02037$05.00